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Could meltwater pulses have been sneaked unnoticed into the deep ocean during the last glacial?

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[1] The lack of climatic imprint left by the Meltwater Pulse-1A ($\simeq$14.5 ka BP), equivalent to a sea-level rise of 14 to 20 meters, is puzzling. Recent studies suggest the event might have occurred as a hyperpycnal flow in the Gulf of Mexico, preventing its detection in oceanic records throughout the North Atlantic. We present a suite of simulations with the LOVECLIM climate model, which mimic the effect of hyperpycnal flow under LGM conditions, in a first attempt to constrain its climatic effects. Analysing the ocean dynamics associated with the anomalous freshwater input, we show that the proposed mechanism is capable of sneaking a significant proportion of the MWP into the ocean ($\simeq$6 meters equivalent sea-level rise using our model under LGM boundary conditions). We also demonstrate that, in our model, the meridional circulation is more sensitive to such inputs in the Arctic Ocean than in the Gulf of Mexico. **Citation:** Roche, D. M., H. Renssen, S. L. Weber, and H. Goosse (2007), Could meltwater pulses have been sneaked unnoticed into the deep ocean during the last glacial?, *Geophys. Res. Lett.*, *34*, L24708, doi:10.1029/2007GL032064.

1. Introduction

[2] Reconstructions show that the changes in global sea-level were not smooth during the last deglaciation (21 to 9 ka BP). Records are marked by events of rapid sea-level rise, suggesting sequences of rapid melting of the remaining ice-sheets. The most prominent one [*Fairbanks*, 1989; *Bard et al.*, 1990, 1996], called the Meltwater Pulse-1A ($\simeq$14.5 ka BP, hereafter MWP-1A) was shown to be a rise on the order of 14 to 20 meters of equivalent sea-level, occurring in 300 to 500 years [*Clark et al.*, 1996]. It should have provided a considerable freshwater flux (FWF) to the global ocean, of about 0.5 Sv (5.10^5 $m^3.s^{-1}$).

[3] Identifying the ice-sheets responsible for such a sea-level change has been a source of debate. Natural candidates are the northern hemisphere ice-sheets which were undergoing strong melting at that time [*Peltier*, 2004]. Among the most likely is the Laurentide ice sheet (LIS) on the basis of evidence [*Flower et al.*, 2004] for low δ^{18}O events recorded in the Gulf of Mexico (GoM) but also being by far the largest contributor to sea-level change during the deglaciation. However, the Antarctic ice sheet has also been suggested [*Clark et al.*, 2002; *Bassett et al.*, 2005] as a possible source for the MWP-1A, although with arguable constraints [*Peltier*, 2005]. Reconstructions show that the Antarctic ice sheet contributed at most 16 to 18 meters to the total last glacial maximum lowstand [*Ritz et al.*, 2001; *Huybrechts*, 2002] arguing against this source as the sole contributor. Assuming that the LIS is the main contributor to the MWP-1A, the problem can be stated as follows: How can an average of 0.35 to 0.4 Sv be added to the Atlantic Ocean during 300 to 500 years without substantially perturbating [*McManus et al.*, 2004] the Atlantic Meridional Overturning Circulation (AMOC) or modifying the water mass properties, so that no drastic low δ^{18}O events are recorded in the northern Atlantic outside the GoM [*Clark et al.*, 1996]?

[4] It is especially puzzling as two deglacial events (Heinrich event 1, HE1 $\simeq$ 17 ka BP, and MWP-1A), occurring one after the other, have a different effect on the AMOC. The iceberg influx associated with the HE1 has left no marked imprint on the sea-level change record but is often suggested to be the cause of a shutdown of the AMOC [*McManus et al.*, 2004]. Conversely, the freshwater perturbation associated with the 20 meters sea-level rise of the MWP-1A has no discernable effect on the same AMOC (although this might depend on the proxy data used to infer the past circulation [see *Robinson et al.*, 2005]).

[5] One plausible reason for this discrepancy is that HE1 occurred in a nearly full glacial state whereas the MWP-1A occurred in the middle of the Bølling warm period [*Stanford et al.*, 2006]. It is also possible that turbulent mixing in the highly baroclinic Gulf Stream (as discussed by *Tarasov and Peltier* [2005]) may have severely diluted low-salinity surface plumes originating from the GoM. Another solution suggested to reconcile the above is to "sneak" the meltwater directly into the deep ocean, hence modifying neither the surface water properties nor the AMOC. It has indeed been shown that a flux of freshwater, when loaded with sediments, could be denser than the oceanic waters in which it enters [*Quadfasel et al.*, 1990] and sink to the bottom of the ocean before loosing its sediment content (a phenomenon named "hyperpycnal flow"). It is known to be a mechanism for renewing deep waters and is observed in the present-day Sulu Sea. The potential for this mechanism to "sneak" meltwater during the MWP-1A has recently been recognized [*Tarasov and Peltier*, 2005] and received support as calcite δ^{18}O records retrieved at different depths in the GoM are showing a mid-depth low δ^{18}O excursion [*Aharon*, 2006, hereinafter referred to as AH06]. This excursion is interpreted as an input of freshwater at the approximate time of the MWP-1A, therefore suggesting a potential explanation for the lack of impact of the dramatic MWP-1A event elsewhere.

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Table 1. Mimicked Hyperpycnal Flow Experiments[a]

Experiment	e.s.l., m	Eq. FP, Sv	FP lo.	ΔT_{CG}, °C	ΔT_{CE}, °C	ΔT_{NA}, °C
gom_12	12.5	0.5	GoM	−6.5	−5	−7
mkr_6	6.3	0.25	MKR	−6.5	−5	−6
gom_6h	6.3	0.25	GoM	−3	−2	−4
gom_6f	6.3	0.25	GoMf	−1.5	−1	−1
mkr_3	3.1	0.125	MKR	−2	−1	−1.5
gom_3	3.1	0.125	GoM	−1.5	−1.5	−3
T&P_10	10.5	N/A	N/A	−2.5	−3.5	−3

[a]e.s.l., eustatic sea level equivalent; Eq. FP, Equivalent Freshwater Pulse. ΔT_{CG} (respectively ΔT_{CE} and ΔT_{NA}) is the maximum temperature anomaly in central Greenland (resp. central Europe and surface North Atlantic (40°N, 30°W)) with respect to the start of the simulation. All pulses have a duration of 300 years. GoMf indicates experiments where the whole GoM is the input area for the FWF. Experiment T&P_10 is with a realistic MWP-1A scenario from *Tarasov and Peltier* [2005, 2006]. See details in Text S1.

[6] Here, we test this hyperpycnal flow mechanism in a global coupled climate model using different scenarios both for the magnitude of FWF and the location of the input in the northern hemisphere. Because we do not seek to reproduce the MWP-1A in full details but aim at understanding whether the hyperpycnal flow mechanism provides an alternative to the classical surface ocean FWF to AMOC relationship for the glacial, we use the LGM as baseline climate for our simulations. Although the LGM climate is not fully comparable with the much warmer MWP-1A climate, it has the advantage of being a well-known equilibrium state in models, a useful reference for our first testing of the hyperpycnal mechanism.

2. Model Experimental Set-Up

[7] In this study, we use the LOVECLIM 3-D coupled climate model [*Driesschaert et al.*, 2007]. Technical description of the model is given in Text S1 of the auxiliary material.[1] The basic LGM climatic state used is described by *Roche et al.* [2007].

[8] As constructing a hydrographic module able to simulate the loading of sediments in river runoff or shelf waters is beyond the scope of this study, we chose to mimic the effect of the hyperpycnal flows by imposing a salinity anomaly in the bottom oceanic cells of the GoM (Figure S1). A salinity anomaly is preferred to a FWF for the sake of simplicity. Because we try to assess if the meltwater could be "sneaked" into the deep ocean, we apply this anomaly from 2 kilometers depth to the bottom ($\simeq$3.5 km depth), a depth quite consistent with what has been observed in the Sulu Sea [*Quadfasel et al.*, 1990]. Moreover, it is consistent with the existence of the Mississippi canyon, which could help the meltwater plume sinking by keeping it coherent inside its borders [*Hallworth et al.*, 1993] until the approximate depth of 1.5 kilometers [*Bryant et al.*, 1991]. We therefore implicitly assume that the sediment loading was sufficient for the flow to be dense enough to reach such depths. Although modern oceanographic observations show that gravity currents can be found at important distances from the source [*Quadfasel et al.*, 1990], it is unclear if the MWP-1A could have filled the whole GoM basin. We therefore defined two different geographical regions for the salinity anomaly: either it fills only half of the deepest part of the GoM (depths below 2 km), or it fills the entire basin at depths below 2 kilometers (Text S1). We also tested different magnitudes for the pulse added into the GoM (Table 1).

[9] As the reactivation of the MacKenzie river (MKR) system has also been suggested as a potential Younger Dryas trigger [*Tarasov and Peltier*, 2005], we tested the mechanism in this region with scenarios where the salinity anomaly is added to the Arctic Ocean following the same methodology. In the following, we focus on the effect of the "hyperpycnal flow hypothesis" in the GoM, but will nonetheless point out the important dynamical differences between the GoM and MKR experiments.

3. Effect of the Mimicked Hyperpycnal Flows

[10] Adding a salinity anomaly in the deep sea (deep GoM or Arctic Ocean) freshens the seawater and thus decreases its density. We chose not to include any anomalous heat content, as the deep waters of the GoM have a temperature of about 1.5°C which is at the extreme lower end of what can be expected from meltwater coming from proglacial lakes at the LIS margin. In the event of a warmer MWP, the effect would even be larger, as warmer temperatures further reduce the water masses' density. The effect simulated here is then an upper end limit of what could be "sneaked" by such processes.

[11] Reducing the densities of the deep waters destabilises the water column, mixing the now less dense deep waters with the overlying ones. Six months after the beginning of the experiment, the whole water column is fresher up to the thermocline (Figure S2). By the end of the first year, the signal reaches the surface of the GoM. Once the surface is reached, the meltwater signal is entrained in the upper branch of the AMOC to the North Atlantic. In case of the MKR experiments, the smaller vertical density contrasts that exists between the deep Arctic Ocean and the surface makes the meltwater influence even larger, and the destabilization process faster. When the anomaly reaches the surface, sea-ice formation is promoted, due to the diminution in Sea Surface Salinity (SSS). The meltwater signal is then advected to the Nordic Seas, both as fresher surface waters and as sea-ice.

[12] δ^{18}O proxy data retrieved in the GoM [AH06] show that the subsurface anomaly is greater than the surface anomaly during hyperpycnal events. Comparing with the time evolution of the simulated salinity anomaly in the GoM

[1]Auxiliary materials are available in the HTML. doi:10.1029/2007GL032064.

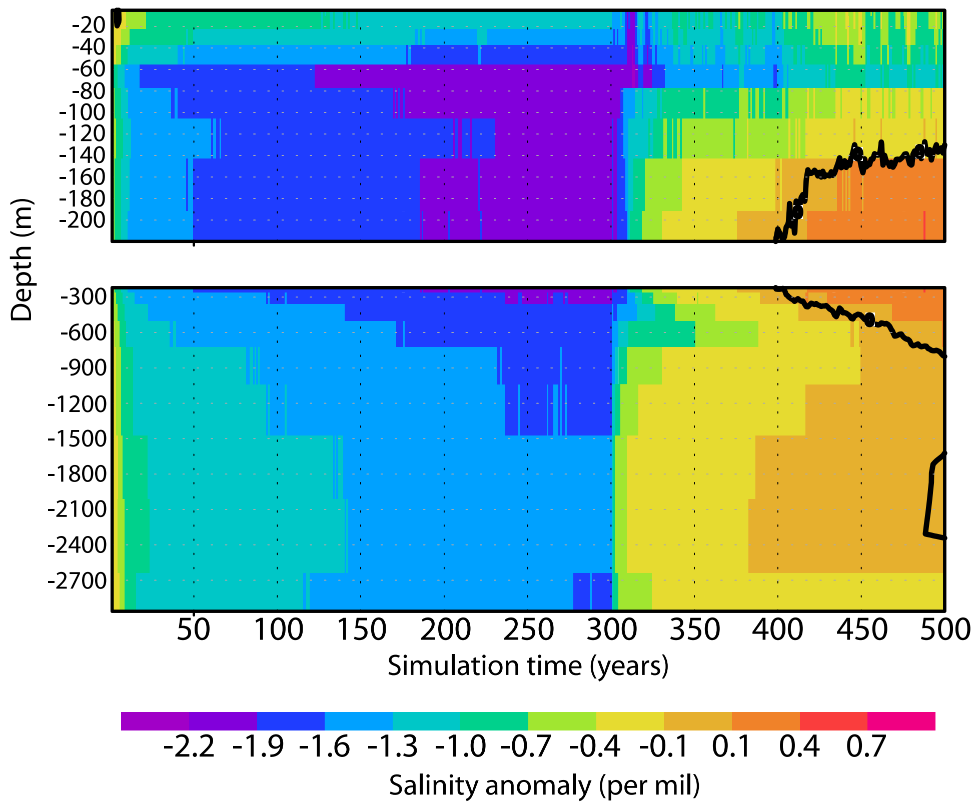

Figure 1. Time evolution of the salinity anomaly with respect to depth in the center of the GoM. The FWF is imposed during the 300 first years of the experiment "gom_3." The thick black line contours the zero anomaly.

(Figure 1), one can see that the maximum salinity anomaly is not at the surface, but between 70 and 400 meters. This result is consistent with proxy data which show a maximum in δ^{18}O between 400 and 530 meters and a bigger anomaly at 100 meters depth than at surface [AH06]. *Flower et al.* [2004] have evaluated the SSS anomaly during the same time period from δ^{18}O proxy data and a simple isotopic model. Although it is difficult to disentangle the strict hyperpycnal period (as defined by AH06, discussion in Text S1) in this study, the proxy data derived SSS anomaly seems to be of about $\simeq$1 per mil, quite consistent with our simulated 0.7 to 1 per mil surface anomaly in Figure 1. Our results for the GoM are therefore in good agreement with available proxy data.

4. North Atlantic Deep Water Formation Weakens in All Experiments

[13] Once advected along the upper branch of the AMOC (GoM experiments), the salinity anomaly eventually reaches the deep water formation sites. Although we use a relatively coarse resolution model, this ought to be the case in any model provided that the upper branch of the AMOC is fed by GoM waters. Not all of the meltwater reaches the convection sites, some of it being mixed via entrainment into the subpolar gyre.

[14] Results obtained under various forcings show different time-dependent behaviour in terms of North Atlantic Deep Water (NADW) export to the South Atlantic (Figure 2). From the initial 16 Sv of simulated deep NADW export, we obtain a reduction in all simulations and a quasi-cessation in two of them ("gom_12" and "mkr_6"). Simulations with GoM input are less sensitive to a given FWF than those with MKR input, and the recovery to a full strength of NADW export is faster after the event. Such differences in dynamic behaviour are due to the location of deep water formation in the modelled initial state. Indeed, two distinct geographical sites are contributing to NADW formation [*Roche et al.*, 2007]: a main one south of Iceland and a secondary, less intense but deeper, in the Greenland-Icelandic-Norwegian (GIN) seas. Adding freshwater in the GoM reduces the activity at the main convection site rapidly, thus promoting considerable reduction of NADW export to the South Atlantic. The GIN Seas site is however less affected, being more remote with respect to the salinity anomaly source. Thus, when the pulse ends, the return of the AMOC to full strength is driven by the GIN Seas convection site. It should be noted that when both convection sites are severely disturbed (e.g. in the "gom_12" experiment), there is no NADW export recovery to full strength in the 500 years we integrate.

[15] Conversely, when freshwater is added in the Arctic ocean (MKR experiments), such as "mkr_6", the GIN Seas site is immediately affected (Figure 2b), but the main convection site, south of Iceland, is more protected, leading to less NADW export reduction at first. After 100 years, the second convection site is also affected, and the NADW reduction is greater at the end of the experiment (e.g., "gom_6h" vs. "mkr_6"). Once the pulse ends, the absence

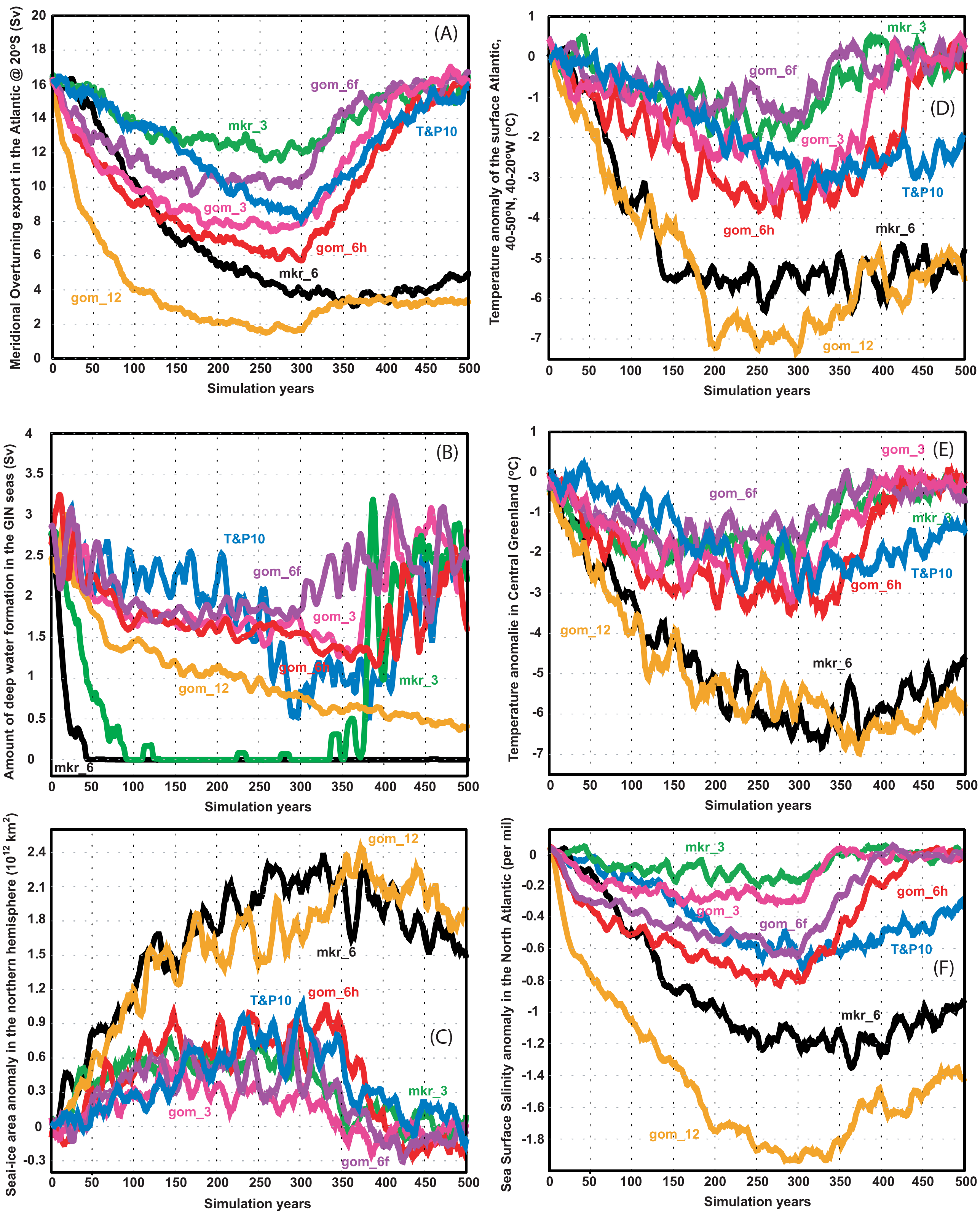

Figure 2. Timeseries for the different hyperpycnal flow experiments. (a) Maximum North Atlantic Deep Water export in the South Atlantic, (b) rate of deep water formation in the GIN Seas, (c) sea-ice area anomaly in the Northern Hemisphere, (d) sea surface temperature anomaly in the North Atlantic, (e) temperature anomaly in central Greenland, and (f) SSS anomaly in the North Atlantic. All series are 10-years running averages.

of convection in the GIN Seas prevents a fast recovery of the NADW export to full strength.

[16] Special mention should be made of the T&P_10 simulation, integrated with reconstructed MWP-1A interval forcings from the LIS [*Tarasov and Peltier*, 2006]. Although the total MWP added is of about 10 meters e.s.l. it is of only 6 meters in the first 300 years. As can be seen from Figure S3, the freshwater added first peaks in the GoM and on the East coast, promoting a strong reduction of the AMOC export to the south Atlantic, mainly affecting the convection site south of Iceland. The following 200 years see a recovering of that site, and an opposite trend at the GIN sea convection site, affected by a relatively strong Arctic Ocean input. The overall response of this more complex T&P_10 scenario is comparable to the gom_6h simulation for the first 300 years, with a slower recovery over the next 200 years, due to the multi-sourced signal.

5. Surface Temperature Response

[17] The simulated reduction in NADW formation limits the northward heat transport in the Atlantic Ocean thereby lowering surface temperatures. Responses from central Greenland and from the north Atlantic indicate that the temperature decrease is linked to the amount of reduction in NADW formation. The relation is however non-linear, with an important role of sea-ice cover as a positive feedback mechanism in further cooling the north Atlantic region. Overall, the temperature response is substantial (-2 to -7°C both in central Greenland and north Atlantic SST) and long-lasting (at least a century). If we consider an arbitrary detection level of 1oC for both temperatures anomalies in central Greenland ice-cores (accounting for inherent uncertainties in measurements (F. Vimeux, personal communication, 2006)). and from deep-sea sediment cores (assuming similar uncertainties), the only experiments that succeed in being "sneaked" unnoticed to the deep ocean are "gom_6f" and "gom_3" (allowing for 3 to 6 meters of equivalent sea-level rise). It should be noted that this level of detection is a lower boundary based on inherent measurements uncertainties only, in an otherwise stable climate. If a transient climate change were imposed, the effective detection level may have been greater. Indeed, it has been shown that during (forced) simulations of the warmer deglacial climate, the freshwater sensitivity of a model could be smaller [*Knorr and Lohmann*, 2003]. It is therefore possible that greater FWF could be "sneaked" in a climatically consistent MWP-1A simulation, the next step to be undertaken.

[18] No simulation with Arctic input would remain undetected in our model, due both to the rapid expansion of sea-ice caused by lowering SSS in the Arctic and its transport to the north Atlantic and by the proximity of the Arctic ocean to the convection sites, limiting the potential for mixing the low salinity signal on the way.

6. Conclusions

[19] While assessing the potential for hyperpycnal flows to "sneak" freshwater into the deep Atlantic ocean, we found that: (1) due to the destabilization of the water column via the imposed salinity anomaly at depth, a meltwater plume still makes its way up to the surface, and is transported to deep water formation sites in the north Atlantic; (2) the efficiency in reducing the AMOC is smaller for freshwater added at depth in the GoM than for freshwater added at depth in the Arctic Ocean; and (3) simulated temperature anomalies are likely to be too strong not to be detected.

[20] We therefore conclude that there is more potential to introduce some freshwater in the ocean as an hyperpycnal flow, at depth, than directly in the upper north Atlantic without drastically altering the climate. In our experiments, we succeed in "sneaking" about 6 meters in the Atlantic ocean, more than what is estimated for HE4 [*Roche et al.*, 2004]. A scenario based on reconstructed LIS drainage chronology showed that using a multi-sourced event slightly increase the ability of "sneaking" freshwater in the ocean, but depend on the precise timing and location. Our results need to be confirmed with other models and tested under full MWP-1A conditions, but already provide a first outlook of the effect of hyperpycnal flow on glacial climate.

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